IMPACTS DO NOT INITIATE VOLCANIC ERUPTIONS. B. A. Ivanov[1,2] and H. J. Melosh[2], [1]Institute for Dynamics of Geospheres, Russian Academy of Science, Leninsky Prospect, 38-6, Moscow, 119334, Russia (baivanov@online.ru, ivanov@lpl.arizona.edu), [2]Lunar and Planetary Laboratory, University of Arizona, Tucson AZ 85721 (jmelosh@lpl.arizona.edu).

Introduction: A theme that runs through many papers on meteorite impact is the idea that large impacts can induce volcanic eruptions through decompression melting of the underlying rocks [eg. 1, 2, 3, 4]. We perform numerical simulations of the impact of an asteroid with a diameter of 20 km striking at 15 km s^{-1} into a target with a near surface temperature gradient of 13 K km^{-1} (“cold” case) or 30 K km^{-1} (“hot” case). The impact ultimately creates a 250 to 300 km diameter crater with approximately 10,000 to 20,000 km^3 of impact melt. However, the crater collapses almost flat and the pressure field returns almost to the initial lithostat. Even an impact this large is insufficient to raise mantle material above the peridotite solidus due to decompression only. The probability that such an impact coincides with the much more frequent occurrence of Large Igneous Provinces seems to be low. We conclude that it is unlikely that a large impact struck such a province any time in post-heavy bombardment Earth history.

Numerical modeling of a large impact event: To illustrate the potential for melting during the excavation and modification of a large impact crater, we performed numerical simulations that include the effects of decompression melting and thermal gradients in the target. For simplicity, we assume a vertical impact by a dunite asteroid into a dunite target. We use dunite because it is described by a reasonably reliable equation of state. Target strength properties are derived from triaxial laboratory tests with an addition of thermal softening - initial cohesion and internal friction decrease gradually as the temperature approaches the melting (solidus) temperature. The simplified melting relationship is used to approximate the near-surface peridotite solidus (figure 3 in [5]).

The numerical simulation uses the Eulerian mode of the SALE hydrocode [6]. Computations reproduce the main phases of a crater growth: transient cavity excavation, gravity-driven collapse with central uplift formation, and a final flat crater. The numerical model resolution (20 cells per projectile radius - 0.5x0.5 km cells) was enough high to make a reliable estimate of impact melt production [7]. The computational zone of a high (0.5 km) resolution covers the area of the transient crater formation (~120 km from the point of impact in horizontal direction and 80 km in depth). Gradual increase in cell size beyond the "high" resolution zone allows us to put rigid computational zone boundaries at ~600 km in horizontal and vertical directions. Initially horizontal rows of Lagrangian tracer particles display the target material displacement.

Results and discussion: Fig. 2 shows the intermediate stage of the transient crater collapse. Tracer layers at the base of the melting zone are shown with thicker lines. Grey shading shows 0-50%, 50-100%, and 100% of melting of rocks at this time moment. This figure illusrtrates that in the "cold" case most deep melted rocks originate from a depth of ~40 km - much less that 125 km nedded for a "trigger volcanism". In the "hot case" melted tocks originate from a depth of ~50 km. However, the base of the melted zone never reach surface and, hence, does not experience the release to zero pressure.

Fig. 3 shows tracer rows displacement in a "hot" case Below the maximum transient crater depth of 50 km, target layers oscillate up and down to a depth of a few km. Layers with an initial depth about equal to the transient cavity maximum depth are involved in the final structural uplift and irreversibly deliver material from depth to near the surface. This material is the most susceptible to pressure release melting. The numerical model yields estimates of the volume of material displaced from one depth interval to another. The model shows that most of melting is due to "normal" shock compression/adiabatic release cycle, enhanced with a higher initial temperature at a depth. Without the shock heating the pressure-relief melting is impossible to this scale of an impact. The total volume of rocks with incipient and complete melting is estimated as ~30,000 km3 and 60,000 km3 for "cold" and "hot" cases correspondingly. The net volume of melt (combining completely melted rocks - above liquidus - and melt from partially melted volumes - between solidus and liquidus) is 2 to 3 times less: 15,000 to 20,000 km3.

An interesting effect of a large scale we find in our "hot" case: the transient cavity collapse "traps" some melted volume as a "neck" around a axis of symmetry (Figure 2, right panel). Once can say that an impact into a "hot" target produces an impact "hot spot". However the geometry of the "hot spot" should be verified with 3D numerical modeling, as 2D modeling often produce artifacts close to the symmetry axis.

What about steeper thermal gradients? A gradient much higher than our “hot” value of 30 K/km already implies the presence of “naturally” melted mantle material close to the surface, making an impact simply a melt excavation event, not a volcanic “trigger”. A region with such a high thermal gradient is already an igneous center whether an impact occurs or not.

The impact of 20 km asteroid may create a deep zone of partially molten mantle material, but *without any significant input from lithostatic pressure release* proposed by many authors. In addition, the probablility to hit a hot spot is 10 to 20 times lower than an impact in a normal "cold" area.

What about much bigger impacts? Consider an impact that creates a transient cavity approximately twice as deep as in our numerical simulation (depth ~100 km). Such an impact is, indeed, big enough to raise hot mantle rocks close to surface. This impact corresponds to a final crater diameter of 400 to 500 km - a very rare event in the current post-heavy bombardment period. Such a huge event is possible, but the lunar cratering record indicates it is highly improbable that an event of this magnitude occurred in the past 3.3 Gyr of terrestrial (and terrestrial planet) geologic history.

Acknowlegements: The work is supported by NASA research grants and Russian Foundation for Basic Science (project # 01-05-64564-à).

References: [1] Ryder G. (1994) *Geol. Soc. Amer Special Paper #293*, p. 11-18. [2] Glickson, A.Y. (1999) *Geology*, *27*, 387-390. [3] Abbott D. H. and Isley A. E. (2002) *EPSL*, *205*, 53-62. [4] Jones, A. P. et al. (2002) *EPSL*, *202*, 551-561. [5] Herzberg C. (1995) In: *Rock Physics and Phase Relations*, *AGU Reference Shelf 3:* Washington, D. C., AGU, p. 166-177. [6] Amsden, A. et al.. (1980). *Los Alamos National Laboratory Report LA-8095*, Los Alamos, NM, 101pp.. [7] Pierazzo et al. (1997) *Icarus*, *127*, 408-432.

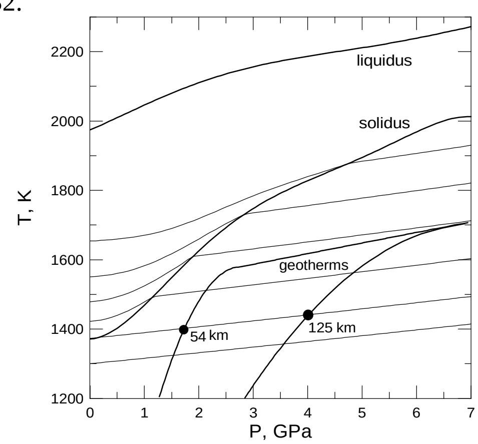

Fig. 1. Simplified pressure-temperature diagram (solidus and liquidus curves) for mantle peridotit adapted from [5]. Adiabats (thin lines) are bended crossing the solidus line where heat of fusion is subtracted in partial melt production. Black dots show crossings of the adiabat of incipient melting at zero pressure with assumed "hot" and "cold" geotherms..

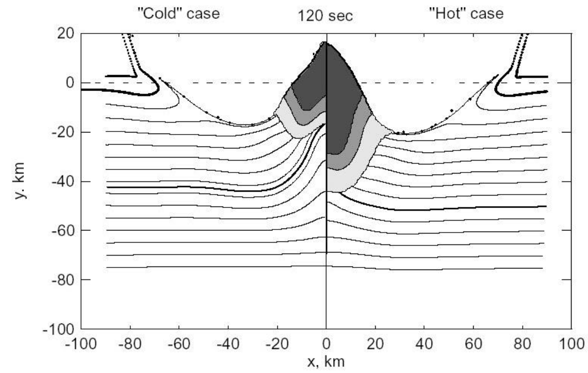

Fig. 2. Displacement of initially horizontal layers of Lagrangian tracer particles in the „cold“ (left panel) and „hot“ target (right panel) at the intermediate stage of transient cavity collapse (120 sec after impact). Thick lines show layers buried initially at the base of a melt zone ~42 km (left panel) and ~55 km (right panel). In the "hot" case (right panel) the transient crater maximum depth of 57 km is comparable to the depth of melting. This results in less effective rebound of the melt zone bottom and to the trapping of a part of molten rocks at a depth (see Figure 3). Gray shading of molten zone corresponds (from light to dark gray) 0% to 50 %, 50% to 100%, and 100% melt content approximately .

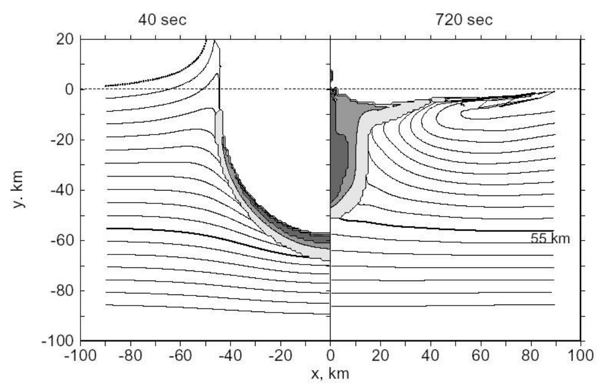

Fig. 3. Displacement of initially horizontal layers of Lagrangian tracer particles in the "hot" target close to a moment of maximum depth of the transient cavity 40 sec after impact (left panel) and at the final stage of cratering when only oscillations of molten material take place. Gray shading corresponds to the same partial melt percentage as in Figure 2. Thick curves show layers buried initially close to the base of a melt zone (~55 km). Mantle at this depth is enough hot to be molten upon release even without a shock heating (see Figure 1). However final uplift of this level is about 5 to 10 km (pressure release 10 to 20% of the initial value) - not enough to reach solidus at the final pressure. Hence melting of mantle at this depth is mostly due to the "normal" shock compression/shock pressure release.